Geomorphology and sedimentology of the Caol Lairig valley, Scottish Highlands: Evidence for local glacier margin advance and retreat during the Loch Lomond Stadial.

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Abstract

A sedimentological investigation of new sections of Loch Lomond Stadial (LLS) age deposits is presented from Caol Lairig valley, located adjacent to Glen Roy, Lochaber, Scottish Highlands. The ice lobes in Caol Lairig and Glen Roy blocked local fluvial drainage systems forming lakes that cut shorelines, the 'Parallel Roads of Glen Roy' (Agassiz, 1840; Jamieson, 1863, 1892). Within Caol Lairig sediment sequences of proximal, distal and deltaic glaciolacustrine sediments and a subglacial till are reported. The till was deposited during ice advance into the valley and the different glaciolacustrine facies formed in the gap between the head of Caol Lairig and the receding ice margin. When the sediments are related to the shoreline and glacial geomorphological evidence, phases of ice advance and ice retreat and the concomitant changes in lake levels are identified. Initially ice retreat in Glen Roy and Caol Lairig was synchronous but after the fall to 325 m the ice in Glen Roy retreated more quickly than in Caol Lairig. Differences in the ice thickness and the lake water depth in Glen Roy and Caol Lairig may have lead to preferential calving of the Glen Roy ice margin hastening ice retreat.

Introduction:

In the UK, the Loch Lomond Stadial (LLS) reflects an abrupt return to cold climate conditions at the end of the Last Glacial Stage (Late Devensian/Weichselian; MIS 2). The LLS is considered to be broadly equivalent to the YD, in Europe, and GS-1 in Greenland, which spanned the period 12.9-11.7 b2k (Rasmussen et al., 2006). The deterioration in climate resulted in the re-advance of glaciers in upland Britain, termed the Loch Lomond Readvance (LLR) (e.g. Clapperton et al., 1975; Sissons, 1979a; Thorp, 1986). These older reconstructions of the ice cap limits (Figure 1A) suggest that, at this scale, lobes of the LLR reached their maxima synchronously and terminated either with distinct terminal moraines or belts of hummocky moraine or within glacial lakes (Rose, 1982; Sissons, 1978; 1979b; 1982; Sutherland, 1984) where the advancing ice blocked local fluvial systems to cause drainage reversals. Recent studies have focussed on annually laminated deposits found within such glaciolacustrine basins (Palmer et al., 2010; MacLeod et al., 2011; 2015), which allow more precise and accurate estimates of the timing and duration of the lake systems. This evidence indicates the LLR reached its maximum toward the end of the LLS in both the north western and southern sectors of the LLR ice cap in Scotland.

Golledge et al. (2008) presented a high resolution numerical simulation of the LLR ice cap that broadly mirrored the known, mapped empirical limits of this ice mass. However, at sites in the northern and southern sectors of the ice cap, where the ice was interacting with the lake waters, the greatest disparity between the modelled ice limits and the empirical evidence exists. For example, in the area around Glen Roy and Glen Spean, there is a substantial overestimation of ice-mass extent by the model in comparison to the known ice limits mapped by Sissons (1979; Figure 1C) in Glen Roy. Consequently, Golledge et al (2008) suggested that further retrodictions of the LLR ice cap would be improved through better understanding of the dynamic nature of ice/lake interactions in the palaeorecord and more refined empirical evidence.

Glen Roy, in the Scottish Highlands, is unique in providing evidence of glacial ice terminating in freshwater lakes for two reasons. First, the ice position in the valley is known through the position of large accumulations of glacigenic deposits in combination with the mapped terminations of the palaeoshorelines in the landscape (Sissons, 1979, c, d). Secondly the deposition of annually laminated sediments enables an estimation of the minimum duration of the lake systems (Palmer et al., 2010; MacLeod et al., 2015; Devine and Palmer, 2016). Thus it is possible to test hypotheses as to the rate of ice advance and ice retreat in Glen Roy, where lake levels rose from 260 m to 325 m and then 350 m as ice encroached into the valley’s mouth. After the ice reached its maximum and retreated, it caused lake level falls from 350 m to 325 m and then to 260 m (Sissons, 1978, 1979b; Figure 1C). These rising and falling limbs of lake development are found to have occurred within a minimum of 515 years (Palmer et al., 2010d).

Whilst this is true of the area where the ice cap reached its maximum in Glen Roy, there is another ice lobe immediately to the west in the valley of Caol Lairig, which also terminated in the Glen Roy glacial lake system (Figure 1B, 1C). This paper presents new sedimentological data from Caol Lairig that can be combined with the geomorphological evidence to establish the position of ice maxima in both valleys. This evidence is used to test the Sissons model of inferred synchroneity of ice advance and retreat during the LLR for both ice lobes, and to provide a detailed account of ice retreat patterns associated with the LLR, particularly for ice terminating in glacial lakes.

Site Context:

Caol Lairig is a tributary valley 1 km to the west of Lower Glen Roy (Figure 1C). It has a relatively small catchment, with a watershed to Glen Roy at an altitude of 297 m OD at the valley's north eastern end. There is high ground c. 700 m to the west forming the watershed to Glen Gloy with small tributary streams, such as Coire Ionndrainn, flowing into Caol Lairig. The river in Caol Lairig passes around the western flanks of Bohuntine Hill (c. 550 m OD) and is confluent with the River Roy after passing through a gorge incised into bedrock to the south of Bohuntine Hill. The bedrock geology of Caol Lairig is dominated by the Brunachan Psammite Formation on the eastern side of the valley and the western flank is underlain by the Ballachulish Limestone Formation (Key et al, 1997)

Sissons (1979) and Peacock and Cornish (1989) have described briefly the geomorphology of Caol Lairig. Sissons (1979) described a terminal moraine at the head of Caol Lairig, which was used to infer the position of the Loch Lomond Readvance maximum. Peacock and Cornish (1989) described a suite of landforms in Caol Lairig formed since the deglaciation of the Late Devensian ice sheet. In Caol Lairig, there are three shorelines eroded into the hillsides at comparable altitudes (260 m, 325 m and 350 m; Figure 1B, 2A, 2C) to those described in Glen Roy (Jamieson, 1863, 1892; Sissons, 1978). In addition, there is a discontinuous shoreline at 334 m present in Glen Roy and on the northern valley side of Caol Lairig, whilst there is a single continuous shoreline at 297 m that is unique to Caol Lairig. A cross-valley ridge, interpreted as a terminal moraine by Sissons (1979b), occurs close to the col between Caol Lairig and Glen Roy at 297 m and is bisected at this altitude at the centre of the valley (Figure 2B). This ridge rises on the northern valley side to merge with a sediment body that rises from 325 m up to 350 m on the northern valley side of Caol Lairig. There are two fan/deltas associated with the 350 m and 325 m shorelines observed in the mouth of Coire Ionndrainn. Peacock and Cornish (1989) also described two shallow sediment exposures in upper Caol Lairig where glaciolacustrine sediment facies provide evidence of a change from proximal to distal glaciolacustrine sedimentation. In one of these sections contorted bedding of these sediment facies was considered as evidence of a palaeoseismic event (Ringrose, 1989), one of two identified in Glen Roy. However the description of the sediments in Caol Lairig is restricted to sections in the northern part of the basin close to the col with little or no detailed description of other sections that are now available within the basin. Also, whilst the geomorphology is considered to relate to glacial advance and retreat during the LLR, it is possible that sediments associated with the Late Devensian ice sheet are present within Caol Lairig. Peacock and Cornish (1989) present the distribution of granitic gneiss in Glen Roy and Caol Lairig, an indicator erratic from the north-west flank of the Great Glen and brought into the Roy catchment when ice flowed over the high hills to the west from the Great Glen during the Late Devensian ice sheet times. One granitic gneiss erratic is observed in Caol Lairig and it might be possible to use this erratic to discriminate between glacial deposits of different ages.

Whereas the sedimentological evidence in Caol Lairig requires further investigation to establish the lithostratigraphy of the basin and is the main purpose of this paper, the geomorphological evidence, in particular the distribution of the shorelines, would suggest that the position of the ice in Glen Roy controls the development of the 350 m, 325 m and 260 m shorelines both in Glen Roy and Caol Lairig. However, the presence of a 297 m shoreline in

Caol Lairig suggests a more complex retreat pattern in this valley than that in Glen Roy. The reasoning for this is that subaerial processes are necessary to act on exposed bedrock and sediment at lake margins for this shoreline to be cut and therefore the lake level in Caol Lairig must be able to fall to 297m controlled by the col at the watershed between Caol Lairig and Glen Roy. Consequently, the lake surface in Glen Roy must have been below the altitude of the col to permit the 297 m lake to exist in Caol Lairig; hence the ice in Glen Roy must have retreated sufficiently to allow the 325 m lake to drain, re-establishing the 260 m lake in lower Glen Roy.

Caol Lairig has clear advantages for improved understanding of events associated with LLR ice advance and retreat. First, there are several thick sediment sections now exposed within a restricted spatial area; secondly the shorelines recorded in Glen Roy extend continuously into Caol Lairig indicating that the control of these lake levels affected both Glen Roy and Caol Lairig and therefore can be linked; and finally, the presence of the unique shoreline at 297 m in Caol Lairig constrains the position of the ice front in Glen Roy. Therefore, this study will combine this geomorphological evidence with the first detailed description of the sediment sequences now available in Caol Lairig. The results lead to refinement of the sequence of events in Caol Lairig and Glen Roy during the Loch Lomond Readvance.

Methods:

Geomorphological mapping of the valley confirmed and added detail to the maps of Peacock and Cornish (1989) by using relief-shaded NEXTMap DSM at a range of scales. Landforms were examined in the field and mapping using 1:10 000 OS maps. Sediment sections were identified with their altitude determined by a Topcon instrumental level. Sediment sections were described using standard lithological techniques and notations (e.g. Evans and Benn, 2004). Clast fabrics were determined from the dip and azimuth of 50 clasts with an a:b axis ratio of 1.5:1 (Kjær and Kruger, 1998). Palaeocurrent measurements were taken from the ripples and imbricated gravels (Tucker, 2003). All fabric data were plotted on equal-area lower hemisphere Schmidt projections. Bulk sediments were collected for particle size analysis by dry and wet sieving for the > 63 um fraction and Sedigraph for the < 63 um fraction following the methods outlined by Gale and Hoare (1991) and Coakley and Syvitski (1991). Clast lithology was undertaken on all units that contained gravel in the range 8-16 mm and 16-32 mm to examine whether different units could be distinguished on their clast lithological compositions. Lithologies referred to as local reflect rock types identified in the immediate catchment of Caol Lairig and consist of psammites, schist (including Leven Schist), quartzite and vein quartz, whereas 'exotic' lithologies refers to clasts from beyond the immediate catchment such as granite, granodiorite, granitic gneiss and igneous rocks undifferentiated. Thin sections of sediment were produced for some of the facies where it was deemed further information was required. Standard thin section preparation techniques were employed to impregnate and produce the thin sections using procedures described by Palmer et al (2008).

Sedimentology and Stratigraphy of the Caol Lairig Valley:

Four main lithofacies associations were observed in six sections in Caol Lairig exposed by river bluffs close to the centre of the valley (Figure 1B, Table 1). Images of the sections and associated lithofacies are presented in Figures 3 to 8. Summary logs of the sections with fabric data are presented in Figure 9, whilst particle size and clast lithological data are summarised in Tables 2 and 3.

Lithofacies 1 is composed of poorly sorted sands and gravels, and diamictons, which vary laterally through the lower part of most sections of Caol Lairig. This unit is divided into two sub-facies.

Lithofacies 1a – Interbedded sands and gravels (Gms; Sh; Sl; Fl).

This is moderately to poorly-sorted gravelly sand (Table 2) with horizontal and crudely bedded, sub horizontal thin beds of sandy gravel. The gravel comprises 85% local lithologies (Table 3) and is dominated by sub-rounded and subangular clasts, which are generally pebble size but also include cobble sizes. The relative proportion of sand and gravel also varies laterally. Toward the top of this lithofacies in section CL 2 there are concentrations of finer sediments with normally graded, laminated fine sands and silts. In the southern part of section CL 2, at the contact to the overlying poorly sorted sand and gravel/diamictons, there is evidence of reverse faulting, dewatering structures and the development of an overturned fold (see below).

Lithofacies 1b - Gravels and Diamictons (Gms; Gh; Sh; Sl; Dmm; Dcm; Dms; Fl).

This lithofacies association is predominantly a matrix supported, poorly-sorted and sub-horizontally bedded gravel (20-49%), with gravel clinoforms dipping at angles between 15-20° toward the NE. The gravels are subrounded and sub-angular, dominated by local lithologies (85-96%) and contain clasts of cobble size but some boulders are also present. Lenses of well sorted, coarse or medium sand that grade upwards into very fine sand and coarse silt are interbedded within the sequences and sometimes display normal or reverse faulting. The diamicton facies of this association is a normally consolidated matrix supported sandy diamicton and are interbedded with the gravels. The a-axis fabric for the clasts in the diamicton is strongly clustered parallel to the main axis of the valley (SW to NE) and the dip of the gravel clinoforms are also aligned parallel to the valley axis. The gravel clast lithologies in the diamicton are dominated by local lithologies. The lateral variation in this lithofacies is best represented in Section CL 2 between 2a and 2b, where a diamicton and poorly sorted gravels at the southern end of the section become more sorted toward the north (upvalley) with the gravels showing steep angled cross bedding (Figure 4).

The massive matrix supported and stratified diamictons are well exposed at the southern end of CL 2 (Figure 4) and directly overlie laminated silts and sands of LFa 1a. Here the contact between LFa 1a and 1b is tilted toward SSW at an angle of 20° in a direction of 90° with the strike 285° to 105°, and the laminated sands and silts form a thrust fold. Toward the apex of the fold nose, beds of gravel are tilted at higher angles (>50 °) and become a greater mix of sand, silt and gravels. Within the laminated sands and silts there is evidence of reverse faults with a throw of 1- 2cm and there is evidence of dewatering structures within the finer grained sediments. The diamictons that overly this sediment facies are succeeded by beds of more sorted gravels or clast-rich diamictons, where the surface of the beds dip at low angles toward the north (upvalley; lower inset Fig 4). In another pit toward the centre of the section of CL 2 similar structures are observed but include a sinistral normal fault allied to the reverse faults with similar throws.

Lithofacies 2 – Diamictons (Dmm)

These are diamicton facies but with distinct properties that allow division into two subfacies. First, Lithofacies 2a (Dmm) is a matrix supported, massive sandy diamicton, which is overconsolidated with a fissile appearance (Figure 3, 7 8). It tends to be a thin unit < 2 m thick with the matrix up to 50% sand. The gravel clasts are dominated by subangular and subrounded clasts predominantly of local lithologies (85-89%) but also it is the only diamicton in the valley that contains granitic gneiss erratics. The strongly clustered a-axis fabric is aligned parallel to the valley axis (SW and NE). At the microscale, the fissility of the sediment is a product of fine, subhorizontal and subvertical channels containing well sorted fine sand, bounded by the diamicton facies. The diamicton has a well-developed, unistrial fabric in places indicating shear and also a masepic fabric suggesting that normal stress has been applied to the sediment.

Lithofacies 2b (Dmm) is a normally consolidated, massive and matrix supported diamicton. This is identified at two points in CL 3; 1) in CL3b where it is interbedded within lithofacies 3, and forms a thin bed (0.4 m) restricted to the upper part of the section CL-3. These are granule to cobble sized, dominated by local lithologies (~88%), subangular and subrounded clasts with a matrix dominated by sand but also a relatively high clay content (25-32%) in comparison to the other diamictons found in the valley. The a-axis of the clasts is moderately clustered and aligned from ESE to WNW. A horizontal gravel unit occurs at the upper contact before passing into laminated sediments of Lithofacies 3; 2) in CL 3b the diamicton retains the same properties as described for 2a but there is a higher proportion of angular and subangular clasts within the matrix and it is thicker (0.8 m) (Figure 5).

Lithofacies 3 Fine grained sorted sands silts and clays (Sm; Fl; Fl(d); Fl(v); Sl; and Sd):

This lithofacies is dominated by fine-grained sediment facies of massive well-sorted sand, deformed sand and laminated very fine sands and silts that are normally graded, but with occasional deformation by folding and faulting. Dropstones of up to cobble size are present within these fine-grained facies. The lithofacies can be further divided into two subfacies: Lithofacies 3a contains sandy and silt laminations with sharp contacts between the laminations. There are infrequent gravel clasts isolated within the lamination structure and also very rare gravel lenses within this facies. Micromorphological analysis of the laminations show that the laminae are normally graded from medium or fine sand to coarse silt with sharp contacts. There is loading at some contacts, but there is little or no clay forming laminae within the sediments. Lithofacies 3b (Fl; Fm; Fld) are well-sorted laminated silts and clays with occasional well-sorted silt lenses. These are horizontally laminated silts and clays with occasional dropstones, but the laminations become increasingly deformed toward the top of this facies and exhibit asymmetrically faulted structures.

Lithofacies 4 Sorted sands and gravels (Gh; GSi, Sfu; St(A); Sr(s)Fl; Fl(d))

This lithofacies is dominated by sands and gravels recorded in the upper part of the sediment sequences toward the head of Caol Lairig (CL4, CL 5 and CL 6; Figure 6, 7, 8). These contain well-sorted, horizontally bedded or laminated silts and sands with climbing ripple structures (Sr A-S), normally graded sands and draped laminations. Horizontally bedded, clast supported gravels are interbedded with the sands toward the top of the sequence at CL 4 with palaeocurrent measurements of the sand facies indicating flow from the NW. The sands and gravels at the top of the unit form part of the large overturned fold. The gravels in these units are dominated by local lithologies and subangular and subrounded clasts.

Position and lateral extent of the Lithofacies associations

All sites examined are positioned to the north and outside of the limits of the 260 m shoreline in Caol Lairig and therefore this lake system. The sections are also all situated within the limit of the 297 m, 325 m and 350 m lake levels defined by the shorelines in Caol Lairig (Fig 1B). CL 1 is the only site that is located within the inferred limit of ice that cut the 325 m shoreline. CL 2 lies 200 m to the north of CL 1 and lies outside the limits of the inferred ice margin when the 325 m lake cut it's shoreline, but within the limits of the ice margin position associated with the 350 m shoreline. All sections are exposed in sections close to the valley axis approximately 200- 300 m from the valley sides except for CL-3, which is 100 m from the eastern flanks of Caol Lairig.

LFa 1a is observed in sections CL2, CL 3a and CL 4. It is relatively thick (up to 2 m), laterally continuous body of sediment observed at the base of these three sections. It is overlain by LFa 1b, which is present in all sections and ranges between 1.5 m thickness in section CL 1 and 6 m thickness in section CL 3b. LFa 2a are thin units of between 0.3 – 1.2 m in section CL 1, CL 5 and CL 6 and usually overlain by LFa 1b. LFa 2b is approximately 0.5 m thick and only observed at the top of section CL 3 situated close to the valley flank. LFa 3 is observed in sections CL 3, 4 and 5, and is between 2 - 4 m thickness in CL3 and CL4, where it is superimposed on the gravels and diamictons of LFa 1b and LFa 2b, whereas in section CL 5 it is observed under LFa 2a. LFa 3a is only present in section CL 3b, and LFa 4 is between 3.5 - 4.5 m thick and is observed at sections CL 4 and CL 6.

Interpretation of Lithofacies Associations:

Lithofacies 1 indicates deposition from a combination of gravity driven transport, high concentration sediment flow processes and also lower energy current flow processes. The horizontally bedded sands and gravels (LFa 1a) show evidence of deposition from turbidity currents with the gravel and sand deposited as part of the traction load (Nemec, 1990; Postma, 1986, 1990). Higher sediment concentrations and flow energy may give rise to more gravel rich facies that are common to Lithofacies 1b. In this instance flows are likely to be dominated by gravity-driven processes closer to the point of sediment release since they are observed on relatively steep angle clinoforms and are interbedded with normally consolidated diamictons, which display clast fabrics that are parallel to flow. Subhorizontal beds of clast- and matrix-supported gravel also exist indicating that some gravel units are deposited as a product of cohesionless debris flows (Mulder and Alexander, 2001). Occasional open framework gravels exist, which suggest that there may have been some remobilisation of gravels across the fan surface probably due to its oversteepened nature. Periods of lower flow regime or sediment concentration are shown by occasional lenses of well sorted and graded sands and silts, which would have formed in small depressions or channels on the gravel surface and may represent lower concentration sediment flows (Lowe 1982).

The relationship between the sediment of Lithofacies 1a and 1b can be observed at the southern end of section CL 2 where the process of diamicton deposition of LFA 1b incorporates finer grained facies of LFa 1a via a thrust fold at the contact between the two units. The nature of the reverse faulting, dykes and thrust fold indicate a lateral compressive force from the SSW with clast rich diamicton partially incorporated into the fold nose. Immediately to the north of this point, farther up valley, the sediments are better sorted with a distinct dip up valley, which suggests that the flows that were transporting these sediments are forced against the valley gradient and are subjected to greater degree of sorting (Hampton, 1972).

This facies is interpreted as a glaciolacustrine proximal subaqueous fan or grounding line fan (Cheel and Rust, 1986; Benn, 1996, Bennett et al., 2002). Flows of high sediment concentration emanate from meltwater channels as cohesionless debris flows and turbidity currents that migrate across the fan surface. These flows can evolve into turbidity currents that give rise to these more sorted and finer deposits while failures on the fan surface can also develop more turbidity currents. The lower energy deposition via suspension settling of Lithofacies 1a may also be formed close to the glacier margin but not directly from the subglacial meltwater channels. The presence of diamictons and glaciotectonised sands and silts, allied to the subangular and subrounded nature of the clasts, implies that ice was in close proximity at the time these subaqueous fan deposits accumulated and occasionally advanced into the sediment stack causing the deformation and dewatering of the sediments, with sediment settling when the ice was in retreat causing the normal faulting.

Lithofacies 2a is an overconsolidated diamicton with subangular and subrounded clasts, a fissile matrix and dewatering structures. Allied to the strongly clustered a-axis fabric it is considered to be a subglacial traction till (Evans et al., 2006). The till exhibits clear indication of deformation caused by both pure and simple shear applied by movement at the bed of a glacier, which caused dewatering and created the fissile, overconsolidated deposit. The fabric data suggests that the till was emplaced directly by a glacier flowing from the south west to north east parallel to the axis of the valley.

Lithofacies 2b has different characteristics to 2a in that it is normally consolidated, does not show the same fissility and has higher clay content than the other diamictons. The fabric data suggest only moderate clustering and also a source from off the valley sides, transverse to the valley axis. These are considered to be either subaerial or subaqueous debris flows deposited down steep, unstable valley sides under gravitational processes (Mulder and Alexander, 2001). In the lower of these two units the clasts are the most angular of the deposits in the valley floor, which reflects the movement of physically weathered material possibly from the eroded shoreline debris to the valley floor. The higher clay content of the sediments shows that distal glaciolacustrine sediments also form the matrix of the sediment and was redistributed to the centre of the basin sediments. The flows developed into turbidity currents with distance into the basin, as indicated by the presence of either massive sands or massive gravels on the surface of the diamicton units.

Lithofacies 3 represents low energy sedimentation on the basin floor of a distal glaciolacustrine context. Sediment delivered to the basin as turbidity currents, which propagates farther into the basin, producing fine-grained sediments (Ashley, 1975; Smith and Ashley, 1985). The lower deposits of LFa 3b are massive or deformed laminations of silts and clays with occasional lenses of sand indicating sedimentation predominantly from suspension with occasional currents of coarser sand-size being deposited (Shaw, 1977; Shaw and Archer, 1979; Johnsen and Brerand, 2006). The lower finer sediments observed in this lithofacies probably represent a more distal sediment source. The grain size decreases from a mode of 8 phi for sediments in CL 3 to 10 phi for CL 4 supporting the view that the principal source of sediment supply was from the south. In addition the presence of deformation structures within the lower finer sediments at both CL 3 and CL 4 indicates an event that was likely to be basin wide, rather than affecting an isolated part of the glaciolacustrine sediments. This may reflect either a palaeoseismic event (Ringrose, 1989) or partial lake drainage. The upper deposits (LFa 3a) are composed of fine sand and silt laminations that are normally graded and draped over dropstones. There is little or no deformation of these laminations. The nature of these sediments suggest that there was episodic delivery of low sediment concentration turbidity currents to the basin with sedimentation again mainly from suspension (Middleton and Hampton, 1976; Nemec, 1990; Postma, 1986, 1990) but with sufficient interval between flows to allow settling of finer grained material through the water column. This part of the sequence also displays evidence of dropstones suggesting ice rafted debris as a source for this material, indicating that ice bergs were being carved from the glacier margin (Thomas and Connell, 1985) and that this is a glaciolacustrine system. This lithofacies is also likely to be the distal equivalent of LFa 1.

Lithofacies 4 is well sorted sand facies interbedded with well and moderately sorted gravels and is interpreted as shallow water deltaic bottomset deposits (Johnsen and Brerand, 2006). The multiple fining upward successions of type A to B to S ripple cross laminations with the sands at the base of the sequence, allied to draped lamination, reflects intermittent sediment inputs (Ashley et al., 1982) attributed to deposition from waning low density turbidity currents (Mulder and Alexander, 2001). Variations in flow velocity and sediment concentration are reflected in the alternation between gravel and sand facies with the gravels deposited via traction and the sands and silts deposited during the waning of the flow (Hampton, 1972). These turbidity currents may also be produced by failures on a delta front. It is noted that the sediment becomes coarser through the succession with silts overlain by sands and gravels which is interpreted to represent the progradation of the delta from finer prodelta facies to coarser facies on the delta front (Bhattacharya and Walker, 1992). Sediment was delivered to the basin as turbidity currents either from failures on the delta foresets or as high-density sediment laden underflows derived from rivers. Paleaoflows indicate sediment sourced from the north-west and probably resulting from newly exposed glacial sediment being reworked via fluvial processes. The subangular and subrounded nature of the gravel clasts suggests that these sediments have not been reworked or transported over any great distance.

In summary, the sediments in Caol Lairig reflect glacial, proximal and distal glaciolacustrine sedimentation within the valley. There is evidence of the redevelopment of a fluvial system in the upper part of Caol Lairig prior to final lake drainage, which probably caused the gully headcutting of these surficial Quaternary deposits (Figure 1B). Further fluvial sedimentation was then concentrated in the course of the existing river channel that was cut during lake drainage.

Discussion:

Intergrating the Sedimentology and Geomorphology of Caol Lairig:

As noted previously the sediments of Caol Lairig have accumulated in a basin where different lake levels existed. CL 3, 4, 5 and 6 occupy positions to the north of the ice limit associated with the 350 m lake level and section CL 2 is just beyond the limits of the ice margin that formed the 325 m lake level. Therefore, LFa 1 represents the formation of a series of proximal subaqueous fan deposits as the ice retreated to the south. Due to the evidence of sediment deformation through a compressive force between the contacts of LFa 1a and 1b toward the base of section CL 2, it is envisaged that this retreat reflected an active, oscillatory ice margin with occasional readvances of the ice front over short distances (Lukas, 2005). The transition from sandier facies (LFa1a) to coarser facies (LFa 1b) also suggests changes in the position of the meltwater channels at the glacier margin since the finer facies of LFa1a are likely to represent sedimentation away from these channels, succeeded by the coarser facies as the product of readvance of the active ice and reorganisation of the subglacial drainage network (Bennet et al., 2002; Kryskowski, 2002; Kryskowski and Zielinski, 2002). Section CL 2 lies ~ 200m within the ice limits for the 350 m lake system and therefore formed after ice retreated from this position presumably during the 325 m lake level stage, whereas the lithofacies in the sections to the north of CL 2 may have been deposited in the 350 m lake. This would indicate that the same processes existed during the 350 m and 325 m lake systems.

The diamicton of lithofacies 2a is interpreted as a subglacial traction till (Evans et al., 2006). It is a relatively thin unit (0.2 -1.2 m thick) present at CL 1, CL 5 and CL 6 such that only thin remnants of this deposit are preserved toward the base of the examined sections. In combination with the moraine that formed at the head of the valley, this deposit supports evidence for Caol Lairig being inundated by ice. The subglacial traction till at CL 1, the most southerly deposit, may be thicker as it was accreting sediment over a longer period of time compared to those in the north of the valley; i.e. during both ice advance and when the glacier margin was receding and lake levels were falling. Also, as the base of this deposit at CL1 is in the riverbed it is not clear whether it overlies any sediments associated with the ice damming during the ice advance phases. The presence of the subglacial traction till in sections either overlying proximal subaqueous fan deposits (CL 6) or finer grained distal glaciolacustrine sediments (CL 5) suggests that deposits associated with the advance phase can be preserved under the subglacial till as a result of the glacier bed sliding over deposits with a high water content (clays) or well drained sediments that do not record a deformation signature such as the gravels. Deposits assigned to lithofacies 2b are one-off subaerial or subaqueous debris flow events only observed close to the eastern margins of the valley toward the top of CL 3.

The sediments of LFa 3 are concentrated in the area around CL3 and CL4. This was probably the deepest part of the basin after ice retreat and the accumulation of the proximal glaciolacustrine sedimentation identified in LFa 1. Here, distal glaciolacustrine sediments (LFA 3a and 3b) may be linked to LFa 4 (only found in CL 4, CL 5 and CL 6 at an altitude of ~ 279 m), forming bottomsets to the delta that was prograding into the basin from the north and northwest, as indicated by the palaeocurrent measurements, mainly from the Allt Coire Ionndrainn, and/or accumulating suspended sediment delivered from the retreating glacier margin to the south. The increase in grain size for LFa 3 in CL 3 perhaps indicates that the source of sediment is becoming more proximal through time. Since LFa 4 also records a similar trend in grain size to LFa 3 it is likely that LFa 4 is a proximal equivalent of the upper part of lithofacies 3. This change in grain size could be achieved through delta progradation into the basin that caused shallowing or alternatively the lake water levels were falling. The examination of the sediment sections have, to date, revealed no sediments interpreted as annually laminated glaciolacustrine sediments.

Within this interpretation of the sediment sequences in Caol Lairig, the best evidence for ice advance is the subglacial traction till (LFa 2a). Subsequently, deposits were laid down during the overall retreat of an oscillating glacier margin. As the ice margin retreated to the south, as indicated by successive reduction of shoreline altitude, sections to the north became more distal to the glacier margin and finer grained glaciolacustrine sediments were deposited. However, a further fall in lake levels caused the northern end of the basin to accumulate deltaic sediments. This explanation is currently favoured because there is limited evidence for ice advance over proximal glaciolacustrine sediments as might be expected if the ice was advancing into a pro-glacial lake, as the Sissons model would suggest. If this was the case a more deformed facies of LFa 1a and 1b overlain by a subglacial traction till would have been expected. This in turn would be overlain by the retreating phase of subaqueous fan deposits with successively finer lake deposits being laid down on the proximal lake facies as described above. It is possible that coarser grained sediments are preserved under subglacial traction till at the northern end of the basin in sections CL 5 and 6, and also below the deposits at CL 1. This could only be tested through borehole and/or seismic work on the sediment sequence.

The large recumbent fold in the gravels of LFa 4 at the top of CL 4 provides evidence of a deformation event. This may have been caused by a number of different factors including a second seismic event identified by Ringrose (1989) or a sudden lowering of the lake waters causing the exposure of supersaturated sediments. Since the section is situated close to the position of the ice margin, seismic activity could be instigated by movements of the ice margin or unloading of the surface crust. The deformation is one of the final events to affect the sediment sequence and therefore the latter two explanations are the most plausible.

Glacial history of Caol Lairig.

The sedimentologiclal data presented enables refinement of the model for the recent glacial history in Caol Lairig to be developed. With the lack of direct dating evidence from these sediments, this model assumes that the deposits in Caol Lairig relate to the LLR rather then the Late Devensian ice sheet on the following basis:

1) The Caol Lairig sediment sequences lie inside the limits of the LLR ice limits in the valley with good evidence of a terminal moraine at the head of Caol Lairig (Sissons, 1979a, d). Therefore the sediments are likely to be the last major sediment accumulations after the retreat of this ice lobe.

2) This terminal moraine is associated with deltas at the margins of the ice limit that are associated with the 325 m and 350 m lake shorelines (Peacock and Cornish, 1989), which suggests a link between the glacier margin and the ice dammed lakes that formed when glacial ice blocked drainage of Glen Roy into Glen Spean. These shorelines have been dated directly in Glen Roy with cosmogenic radionuclide ages (Fabel et al, 2010) of 11.5 ka +/- 1.2, whilst glaciolacustrine varve sediments in Glen Spean and Glen Roy have two Katla type cryptotephra correlated to the Vedde Ash and Abernethy Tephra (MacLeod et al, 2015) dated to 12,121 cal yrs BP and 11.72 – 11.23 cal yrs BP respectively. Both independent dating techniques therefore support the concept of LLR ice in Caol Lairig.

3) clast lithological analysis of the gravels and diamictons in Caol Lairig reveal no substantial differences in the clast lithology between different units, which are dominated by local lithologies. The presence of a limited number of granitic gneiss clasts in the subglacial till may have resulted from erratics being deposited in glacial sediments in Caol Lairig during the Late Devensian ice sheet times (c.f Peacock and Cornish, 1989), and subsequently reworked by the Loch Lomond Readvance glacier, which transported and deposited mainly local lithologies.

The following phases combines the sedimentology of this study and the geomorphological evidence from Peacock and Cornish (1989) to produce a model for the evolution of the lake basins in Caol Lairig and Glen Roy. Figure 10 presents a plan view of the lake extent in Caol Lairig and the area around the Viewpoint of Glen Roy, as well as a cross section of the lake levels and position of the ice margin in Caol Lairig.

Phase 1: Ice advances up Caol Lairig first into the 297 m lake held up against the col into Glen Roy to form glaciolacustrine deposits. Ice also advances into Glen Roy and as the River Roy is blocked from draining into Glen Spean the waters rise in Glen Roy to 325 m and later inundate Caol Lairig. These glaciolacustrine sediments are overridden by ice and deformed as subglacial traction till is deposited at CL 1, CL5 and CL 6.

Phase 2: Ice Maxima. The terminal moraine at the head of the Caol Lairig is emplaced as the ice reaches its maximum limit. Ice remains here to form ice marginal landforms, such as the terminal moraine at the head of Caol Lairig and either deltas or kame terraces, which form at 350 m on the northern flank of Caol Lairig. This indicates the ice in Glen Roy was at or close to its maximum beyond the Viewpoint at this time. The ice advance causes the removal of most sediment from Caol Lairig, although at the base of CL5 laminated fine sediments shows that earlier phases of the lake system may be preserved under the current exposures of sediment. The ice advance possibly erodes the 260 m, 325 m and 350 m shorelines that formed during the rising sequence in Glen Roy.

Phase 3: Ice Retreat: The development of negative mass balance is sufficient for the ice in Caol Lairig to retreat. This forms a shallow basin between the moraine at ~ 297 m OD and the ice margin. With increased meltwater availability, high sediment loads are released from the glacier meltwater portals to form proximal subaqueous fan deposits. Where ice is in contact with the fan diamictons or poorly sorted sands and gravels of LFa 1 were deposited and in more distal locations finer grained sediments accumulated. Aggradation of the subaqueous fans initially occurred in the 350 m lake system with the thickest accumulation of deposits (~8m) occurring in the vicinity of CL2 and 3. At CL 2, compressive deformation of sand and gravel beds implies that the ice margin was advancing on occassions.

Phase 4: Ice retreat continues with lake levels falling in Glen Roy to 325 m when the col through Gleann Glas Dhoire opened. The position of the ice margin at 325 m is clearly defined on the basis of the southerly termination of the shorelines in Caol Lairig, where the ice only retreats ~ 300 m to the south. Proximal glaciolacustrine sediments continue to form at the glacier margin and the formation of the initially finer grained distal glaciolacustrine sediment continues in ~ 50 m of water depth. It should be noted that the ice margin in Glen Roy was at least 2 km wide and the water depth ~ 200 m whereas in Caol Lairig the ice front was a minimum of 0.75 km wide and only 80 m deep.

Phase 5: The further retreat of the ice in Caol Lairig enables the formation of the 297 m lake, which is restricted to Caol Lairig valley. The evidence for a lake at 297 m to form in Caol Lairig is the shoreline, which can only form under subaerial processes with freeze-thaw dominating. Therefore, in Glen Roy the lake water must have fallen to 260 m to expose the col at 297 m OD at the head of Caol Lairig that controls this lake level. This situation would occur when lake water escaped through the mouth of Glen Roy and into the Spean valley to drain through the Pattack - Mashie col at the eastern end of Loch Laggan. By this point, the ice in Glen Roy had retreated by ~ 6 km, whereas in Caol Lairig the ice had retreated only 2 km. Also, the shallow proximal deltaic sediments of LFa 4 in CL 4, CL 5 and CL 6 were deposited via streams emanating from Allt Coire Ionndrainn and the coarser distal glaciolacustrine facies present at the top of sections in CL 3 with some sediment coming from the north west and dropstones being deposited from the icebergs as the ice margin calved into the shallow lake.

Phase 6:
The ice retreats or thins allowing the 297 m lake in Caol Lairig to drain into the Roy to the south of Bohuntine Hill and establish the 260 m lake in Caol Lairig. The presence of the 260 m shoreline in Glen Spean suggests that this shoreline continued to develop as the ice retreated farther down the Spean valley toward Spean Bridge until the final drainage of the 260 m lake.

Implications:
The precise configuration of ice recession in Caol Lairig is more complicated than originally thought. Rather than the linear advance and retreat proposed by Sissons (1979) with predictable recession from 350 m to 325 m and then 250 m, there appears to be faster retreat of the ice lobe in Glen Roy than in Caol Lairig. The evidence presented here, and that builds on the work of Sissons (1979b) and Peacock and Cornish (1989), shows that the glacier in Caol Lairig retreated initially at the same rate as in Glen Roy. Ice, after reaching its maximum in Glen Roy, retreated but maintained a competent barrier to retain the lake waters at 350 m. This explains the presence of a 1.5 km stretch of 350 m shoreline inside the limits of the maximum ice position to the north of the Viewpoint. Likewise in Caol Lairig, ice retreat of 1.4 km can be inferred from the presence of the 350 m shoreline to the south of the terminal moraine at the head of Caol Lairig. The varve chronology from Lochaber is used to suggest that the minimum duration for the glacial lakes is 515 years (Palmer et al., 2010; Devine and Palmer, 2016) and that the retreat phase (350 to 260 m lake levels) of this was ~ 200 years. Therefore 100 years would have elapsed during the 350 to 325 m stages implying rates of retreat of 14 and 15 ma^{-1} for Caol Lairig and Glen Roy respectively.

It is still unclear how initial glacier retreat in Glen Roy allowed drainage through the 325 m col at the head of Gleann Glas Dhoire. There is an intermediate shoreline at 334 m in Caol Lairig (on the northern valley sides Figure 1B); a similar one occurs in the vicinity of the Glen Turret Fan (on the northern valley sides) and by Bohaskey (north east/east valley sides). It is possible that, for a short period, 350 m lake water drained over the surface of the ice or around its margin at an altitude of 334 m into the 325 m ice-marginal lake in Gleann Glas Dhoire but with sufficient time for short sections of shoreline to be cut in susceptible bedrock at specific locations in the valley i.e. those with a southwest or westerly aspect and with the greatest fetch. The evidence presented here suggest that ice in Caol Lairig and Glen Roy retreated actively over similar distances during this period and therefore favours the argument for synchronous retreat of the two ice lobes rather than ice thinning.

Later, the rate of ice retreat decreased in Caol Lairig compared to Glen Roy. A ~200 m stretch of 325m shoreline exists in Caol Lairig beyond the southern terminus of the 350 m shoreline. Whereas in Glen Roy another 1.5 km segment of 325 m shoreline is revealed, the southern end of which is linked to another moraine ridge around Bohuntine Hill suggesting a glacier stillstand. The ice in Caol Lairig was receding slowly in comparison to Glen Roy, which was receding at a similar pace to the initial retreat of the 350 m lake system.

At this stage, the role of lake water depth in relation to the thickness of the retreating ice margin in Glen Roy and Caol Lairig may explain this disparity. For example, in Caol Lairig the ice front forming the 325 m lake level is terminating in 65 m of freshwater with an unknown ice thickness. In Glen Roy at the same lake level, ~195 m of lake water is dammed against an ice mass of between 325 m and 334 m altitude. Water depth is nearly 3 times greater in Glen Roy than in Caol Lairig, which, if assuming that calving was a major form of ice loss at the margin, would cause calving speeds three times as great in Glen Roy than in Caol Lairig (Funk and Röthlisberger, 1989). This would encourage greater draw down of ice into the lower valley of Glen Roy and cause thinning of the ice mass, which might reach a critical level for water to flow over the ice surface toward the Spean. This may partly explain the evidence for greater movement of the ice front in Glen Roy than in Caol Lairig during this phase of retreat. However, the topographic setting, the role of thermal undercutting and whether the ice became buoyant need to be considered further before firm conclusions can be drawn on the nature of ice retreat.

It is difficult to assess this relative competence of the ice lobes for the subsequent lake levels and mechanisms for the drainage of the lake systems. When there is evidence for the 297 m lake level in Caol Lairig, the lake waters must have fallen to 260 m in Glen Roy. In Caol Lairig the ice limit for the 297 m is only 400 m to the south of the 325 m lake level ice limit, but in Glen Roy it is difficult to pinpoint the location of the ice front was when the lake levels dropped to 260m. This is because there are remnants of 260 m shoreline preserved along the length of lower Glen Spean. It is only possible to infer that once the lake level had fallen to 260 m then the ice in Caol Lairig had sufficient competence to allow a lake to persist and a shoreline to be eroded at 297 m prior to ice retreat in Glen Roy and the re-unification of the Caol Lairig and Glen Roy lake systems at the 260 m level. Nonetheless, the initial phases of ice retreat are recorded well within these sequences.

The information presented here could be used in developing the parameters for retrodictive glacier models. Establishing the position of the ice limit, whether the ice was grounded at its maximum limit, the velocity of the ice at the ice front and knowing the volume of water that the ice is damming, it should be possible to test if the threshold for the ice front to become buoyant in the valley and when the glacier loses its capacity as an ice dam occurs. It should also be able possible to use this glacier behaviour to refine the model iterations of Golledge et al (2008), in particular where the glaciers are terminating in freshwater lakes, to establish the forcing factors behind the development of the LLR ice cap.

Also of significance are the two deformed horizons in the lake sediments of lithofacies 3 and 4. The first in laminated silts and clays of a distal glaciolacustrine deposits (LFa 3b) relate to either the fall in lake level from 350 to 325 m or from 325 m to 297 m. The second deformation event occurs in shallow deltaic sediments (lithofacies 4) at an altitude of ~270 m and probably relate to the drainage of lake waters from 297 m to 260 m. It is not clear whether these defomed beds in Caol Lairig can be attributed to palaeoseismic events or one-off surge type events. If this is a palaeoseismic event in Caol Lairig then this is in addition to the two already identified by Ringrose (1989) and one of which is unrelated to those recorded in Glen Roy, since this latter lake level drop does not occur in Glen Roy. The origin of these deformation events requires further detailed examination of the lithostratigraphy of the deposits in Glen Roy around the Viewpoint and Middle Glen Roy.

Summary:

- New sediment sections in Caol Lairig have been described and, through the relationship to the landform evidence in the valley, these sediments can be attributed to the advance and retreat of an small ice lobe during the Loch Lomond Readvance.
- Sediments in Caol Lairig display subglacial tills formed during ice advance and proximal grounding line fan deposits, distal glaciolacustrine laminated and deltaic deposits, which formed as the ice retreated and forming an ice contact lake system connected to Glen Roy.
- No glaciolacustrine varve sediments have been discovered in Caol Lairig.
- The lake water level in Glen Roy is controlled by the position of the ice, which also controls the lake levels in Caol Lairig..
- The presence of a single intermediate shoreline unique to Caol Lairig at 297 m suggests that ice retreat in Glen Roy was quicker than in Caol Lairig before the final demise of the 260 m lake system.
- Deeper lake waters in Glen Roy may have contributed to this faster ice retreat rate than in Caol Lairig.

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Figure Captions:

Figure 1: A) Location of the study area in Scotland on the north-western side of the Loch Lomond Readvance ice cap (after Golledge et al, 2008) B) Geomorphological map of Caol Lairig adapted from Peacock and Cornish (1989) and additional information from NEXTMap DTM (courtesy of Intermap Technologies Ltd) and reconnaissance fieldwork during the analysis of the sediments for this study. C) Sissons model for advance and retreat of ice into Glen Roy and Caol Lairig with the associated rise in lake levels controlled by the position of the ice in Glen Roy. Note that ice in Glen Roy reaches its maximum synchronously with Caol Lairig and the ice margin when in retreat is assumed to reach the same positions as during the advance phases. Plotted using Digimap Ordnance Survey data, Crown Copyright (2015).

Figure 2:

A): View of northern Caol Lairig from the western flank of Bohuntine Hill. The 350 m and 325 m shoreline are visible in northern Caol Lairig and extend into middle Glen Roy. The northern part of moraine at the head of CL is with ice marginal fans associated with the 325 m and 350 m shorelines visible. In the bottom left of the photograph the section CL 4 can be seen exposed in the valley floor. B): Detail of the moraine at the head of Caol Lairig with a view looking toward the south and Bohuntine Hill, where the moraine appears to run obliquely up the valley side. The 350 m and 325 m shorelines are visible on Bohuntine Hill and extend beyond the moraine toward the watershed to Caol Lairig. The ice proximal face of the moraine is to the right and the valley floor where the moraine is dissected is at an altitude of ~ 297 m and would have formed the control height for the independent lake in Caol Lairig of 297 m. C). Image of Caol Lairig viewed toward the south with the western flank of Bohuntine Hill where the 350 m, 325 m and 297 m shorelines are visible. In the valley floor section CL 3 has been formed by erosion from the river in Caol Lairig. The section height is ~ 12 m high.

Figure 3: Site CL 1 with LFA 2a diamict exposed at the base of the section and overlain by gravels of LFA 1b. Spade for scale.

Figure 4: Centrepiece: Section CL 2 exposed in the western bank of the stream with the upstream toward the head of the valley to the right. The gravelly sand facies of LFA 1a is exposed in pits at the base of the section and is overlain by a sharp erosional contact of the gravel facies of LFA 1b. Top left image provides an example of the cross bedded gravel facies within LFA 1b which overlie the diamicton n the lower left hand corner of the image. Top centre; finer grained facies in LFA 1b deposited in a small channel eroded into the gravel. Note the reverse fault within the bed. Top right: Detail of the sand lens in previous image with sands with dune structures at the contact overlain by draped silt bed. Bottom left: Southern end of section CL 2 with the thrust fold of LFA 1a material combined with the diamicton of LFA 2a. Bottom centre: Detail of the contact between laminated sands and silts at the top of lithofacies 1a, which display draped lamination, reverse faults, dewatering structures and flame structues, but also the sediment has been tilted subvertically. An annotated figure is provided to give examples of the reverse faults. Bottom right: overview of LFA 1a; the gravelly sand facies observed at the base of the section. Trowels or spades are used for scale,

Figure 5: Top left; LFA 1a at the base of CL 3 overlain by LFA 1b; Top centre; CL 3 detail at the contact between LFA 1a and 1b with reverse faulted gravelly sands overlain by the poorly sorted sandy gravel of LFA 1b; Top right; detail of LFA 1b poorly sorted massive gravels in CL 3; Bottom left; lower clay dominated laminated facies of LFA 3, with the colour differences showing the increased sand content in the deformed laminations toward the top of the sediment unit. Bottom centre: LFA 3 subaqueous debris flow deposit formed of diamicton at the base but with increased clast content in the upper part of the unit. There is a sharp contact to the overlying silt and sand laminations of the upper part of the lithofacies. Bottom right: normally graded silt and sand laminations draping a clast protruding from the underlying subaqueous debris flow.

Figure 6: Top left: Overview of section CL4; Top right; LFA 1b observed in the lower part of the section. Bottom left; lower unit of Lithofacies 3 observed in

the middle part of section CL 4. Bottom right; folded sands and gravels of lithofacies 4 in CL 4.

Figure 7: Section CL 5 with the fine grained laminated sediemtns at the base of the unit overlain by the overconsolidated till of LFA 2a, which is succeeded by LFA 1b.

Figure 8: Section CL 6 with sandy gravels of LFA 1b, overlain by LFA 2a and a thin veneer of LFA 1b. Laminated silts and sands (LFA 4) cap the sequence to the top of the section.

Figure 9: Summary lithological logs of the six sections described in Caol Lairig. Lithofacies codes are taken from Evans and Benn (2004). All section depths are provided on the y-axis with figures in brackets referring to the OD height for the top and bottom of each section. A-axis fabric data and palaeocurrent information is presented in the bottom right corner for the key lithofacies coarse-grained lithofacies.

Figure 10: Schemmatic model for the ice advance and retreat in Caol Lairig and Glen Roy. The diagrams include a plan view of the ice extent and lake extent during the LLR in both Caol Lairig and Glen Roy, whilst the cross sections illustrate the likely build up of the sediment succession in Caol Lairig, and the possible lake levels altitudes.

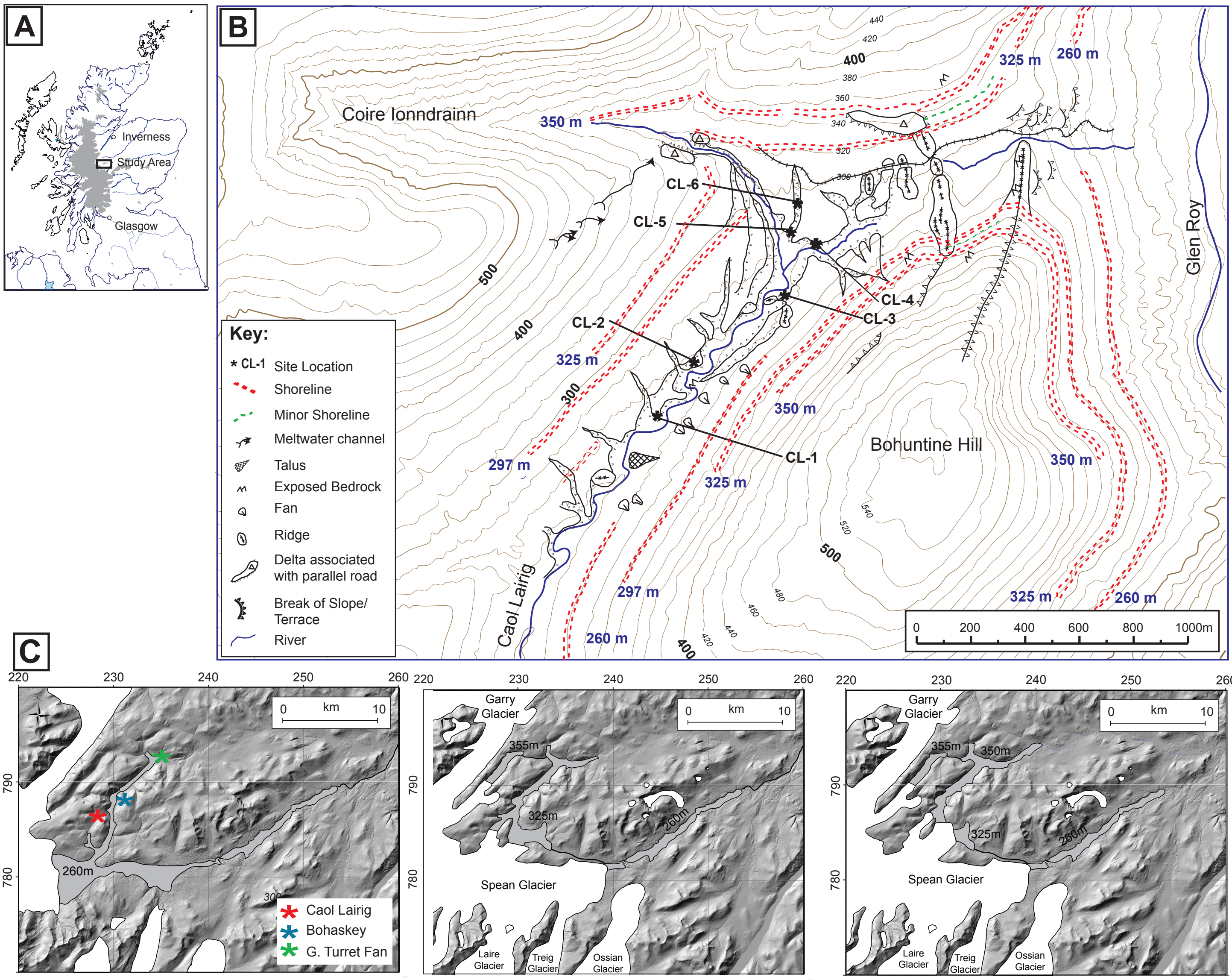
Table Captions:

Table 1: Summary of location, altitude, and lithofacies observed at each sediment section examined in this study.

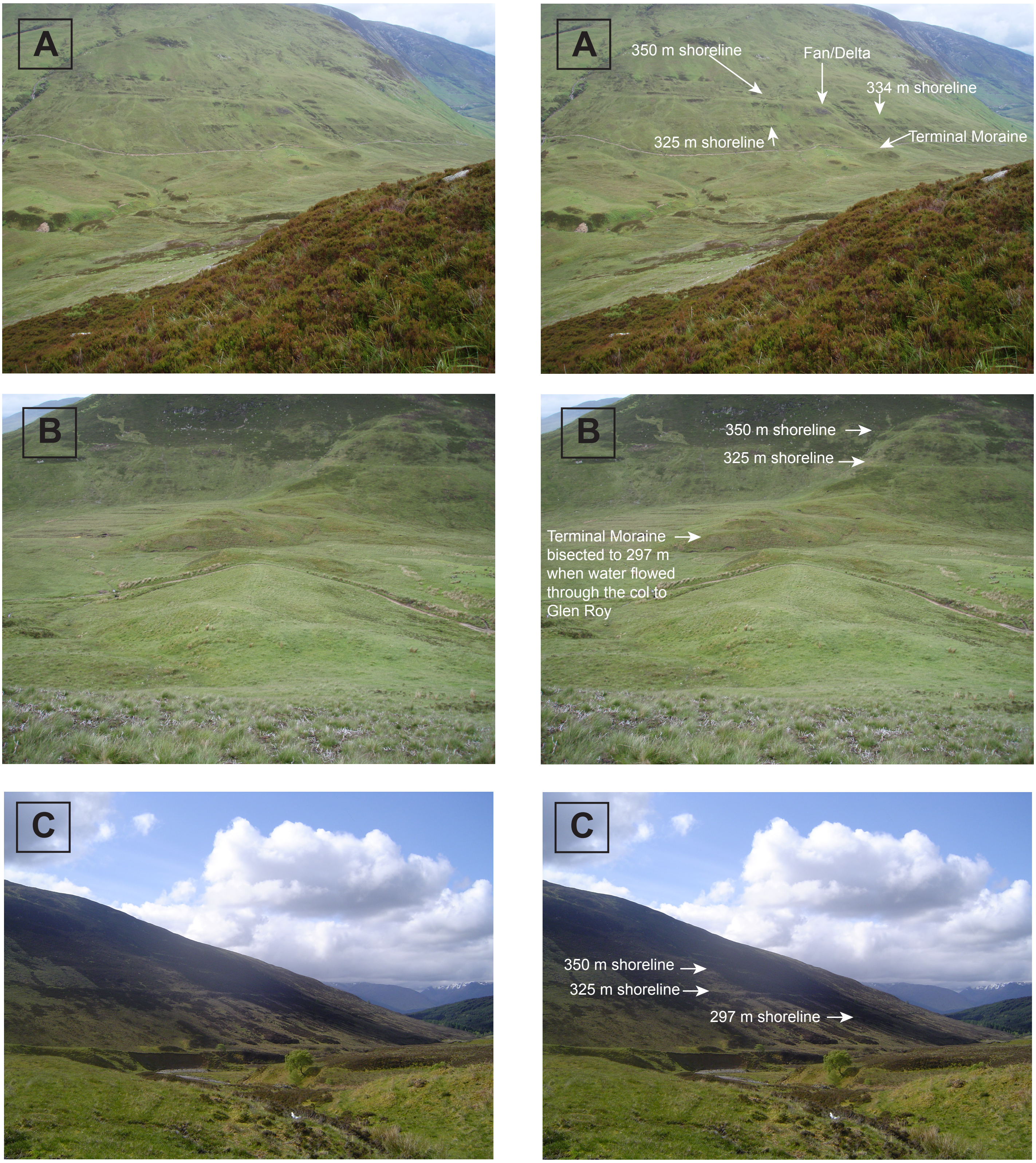
Table 2: Summary table of grain size analyses as examples of the different lithofacies 1-4. Grain size was measured on the <4 mm fraction and therefore the sand to gravel ratio may differ from that quoted in the main text.

Table 3: Summary table of examples of the clast lithological analysis of the gravel units in Caol Lairig. All samples are dominated by local lithologies except for the subglacial traction till of LFA 2a, which has a high proportion of undifferentiated lithologies when compared to the other units.

Figure 1



A
Inverness
Study Area
Glasgow
B
Coire Ionndrainn
350 m
325 m
260 m
400
CL-6
CL-5
CL-4
CL-3
CL-2
CL-1
325 m
297 m
350 m
325 m
Bohuntine Hill
350 m
297 m
260 m
325 m
260 m
Caol Lairig
Glen Roy
500
400
300
500
Key:
CL-1 Site Location
Shoreline
Minor Shoreline
Meltwater channel
Talus
Exposed Bedrock
Fan
Ridge
Delta associated with parallel road
Break of Slope/ Terrace
River
0 200 400 600 800 1000m
C
260m
Caol Lairig
Bohaskey
G. Turret Fan
0 km 10
Garry Glacier
355m
325m
260m
Spean Glacier
Laire Glacier
Treig Glacier
Ossian Glacier
350m



Meltwater channel

Figure 2

Figure 3

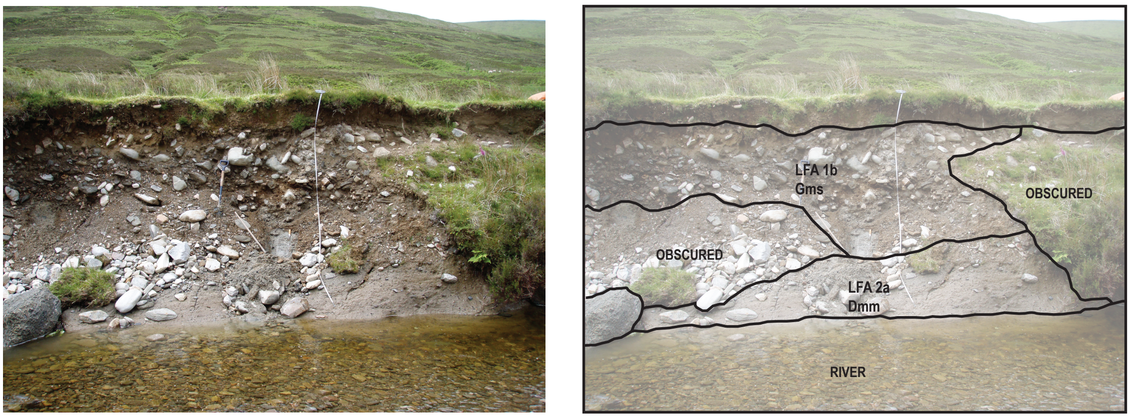
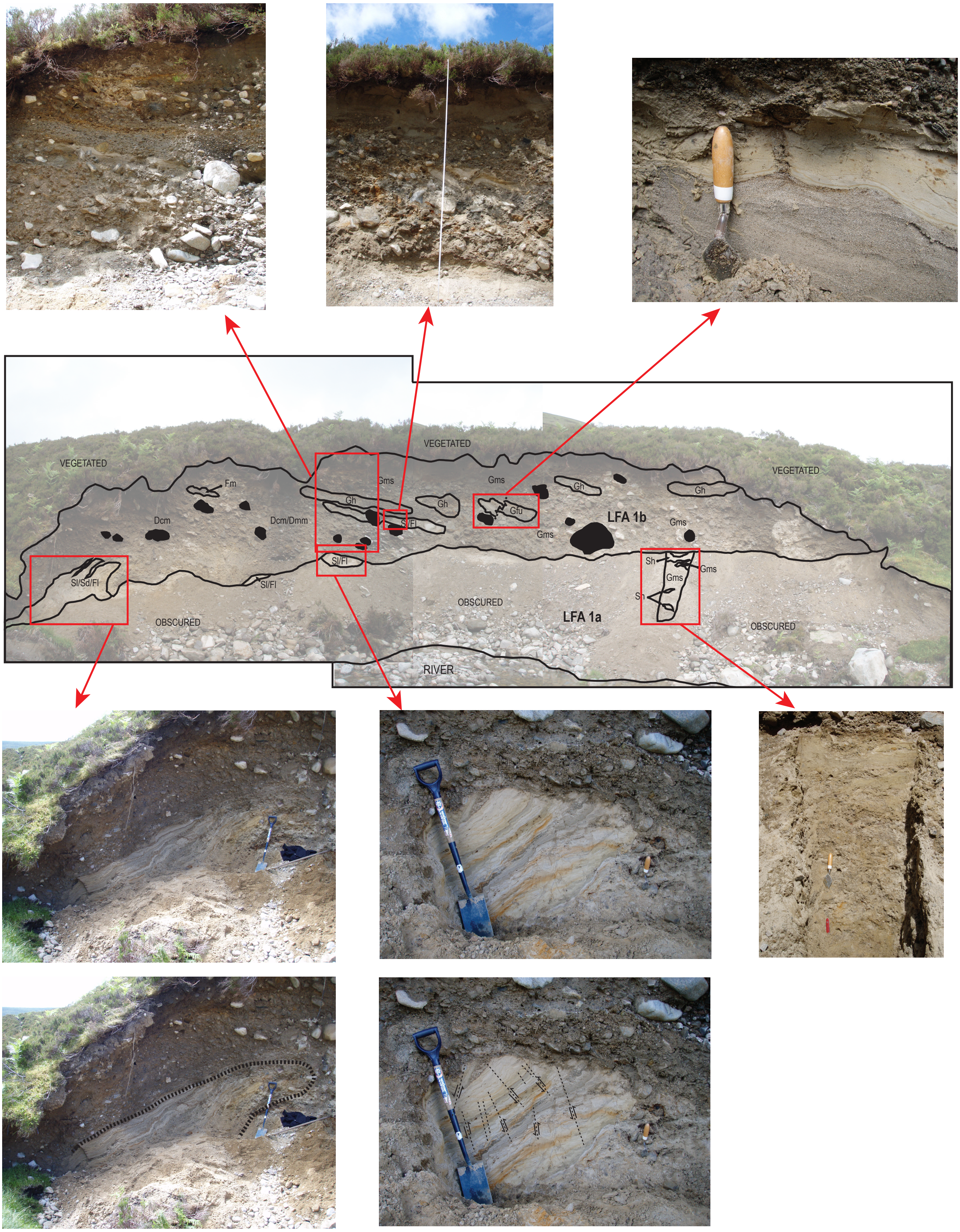


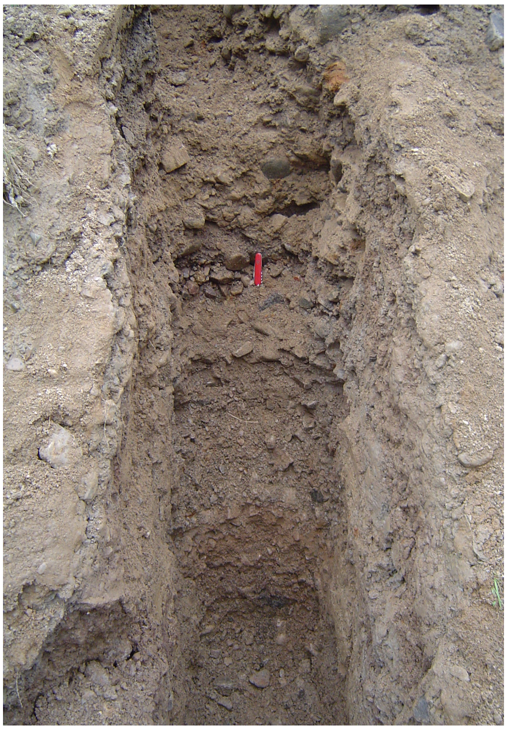
Figure 4



VEGETATED
VEGETATED
VEGETATED
Fm
Dcm
Dcm/Dmm
Gms
Gh
Gh
Sl/Fl
Gms
Gfu
Gh
Gh
LFA 1b
Gms
Gms
Sl/Fl
Sl/Sd/Fl
Sl/Fl
Sh
Gms
Gms
Sh
OBSCURED
OBSCURED
LFA 1a
OBSCURED
RIVER

Figure 5

Figure 6



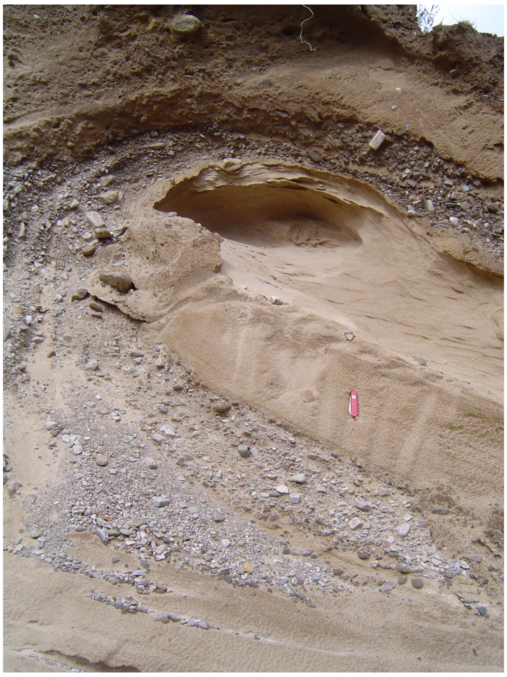


Figure 7

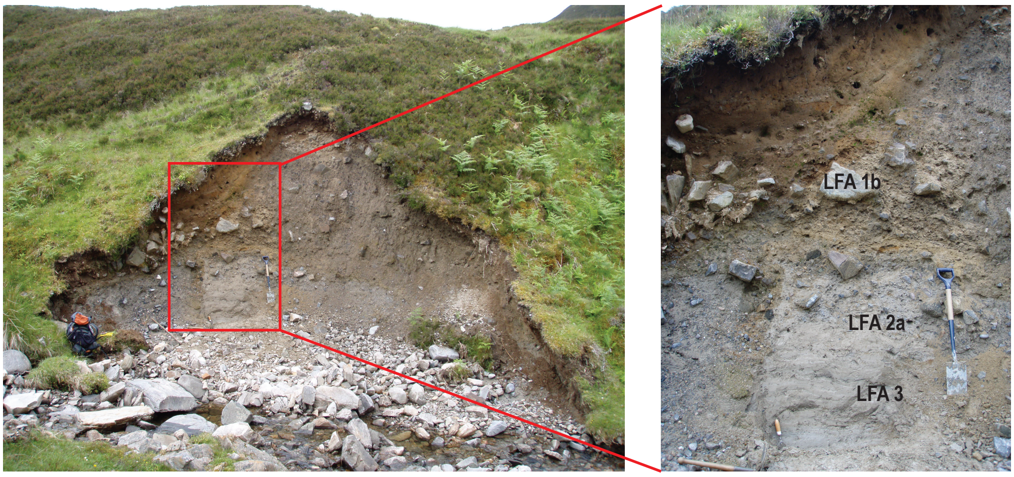


Figure 8

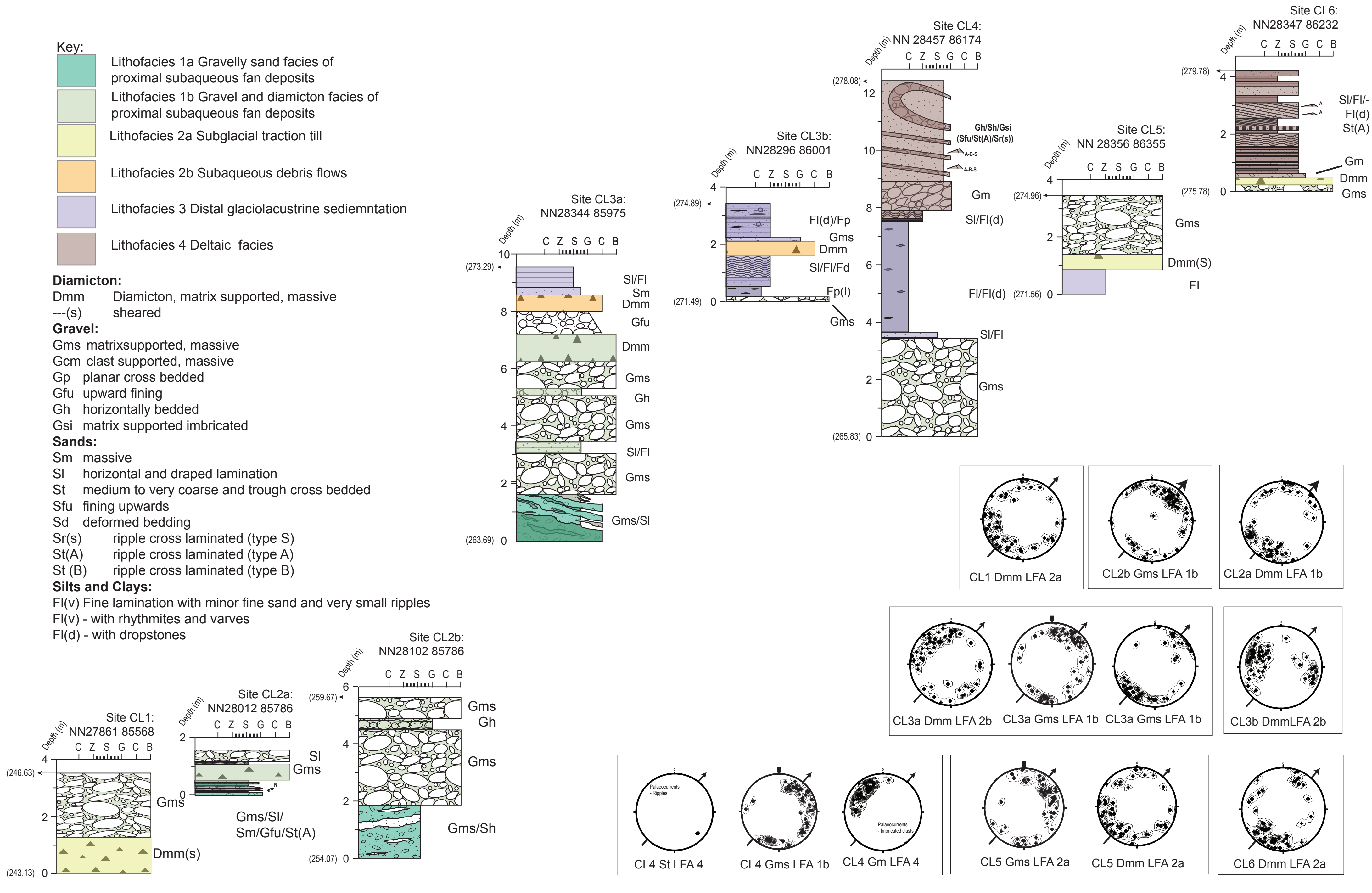
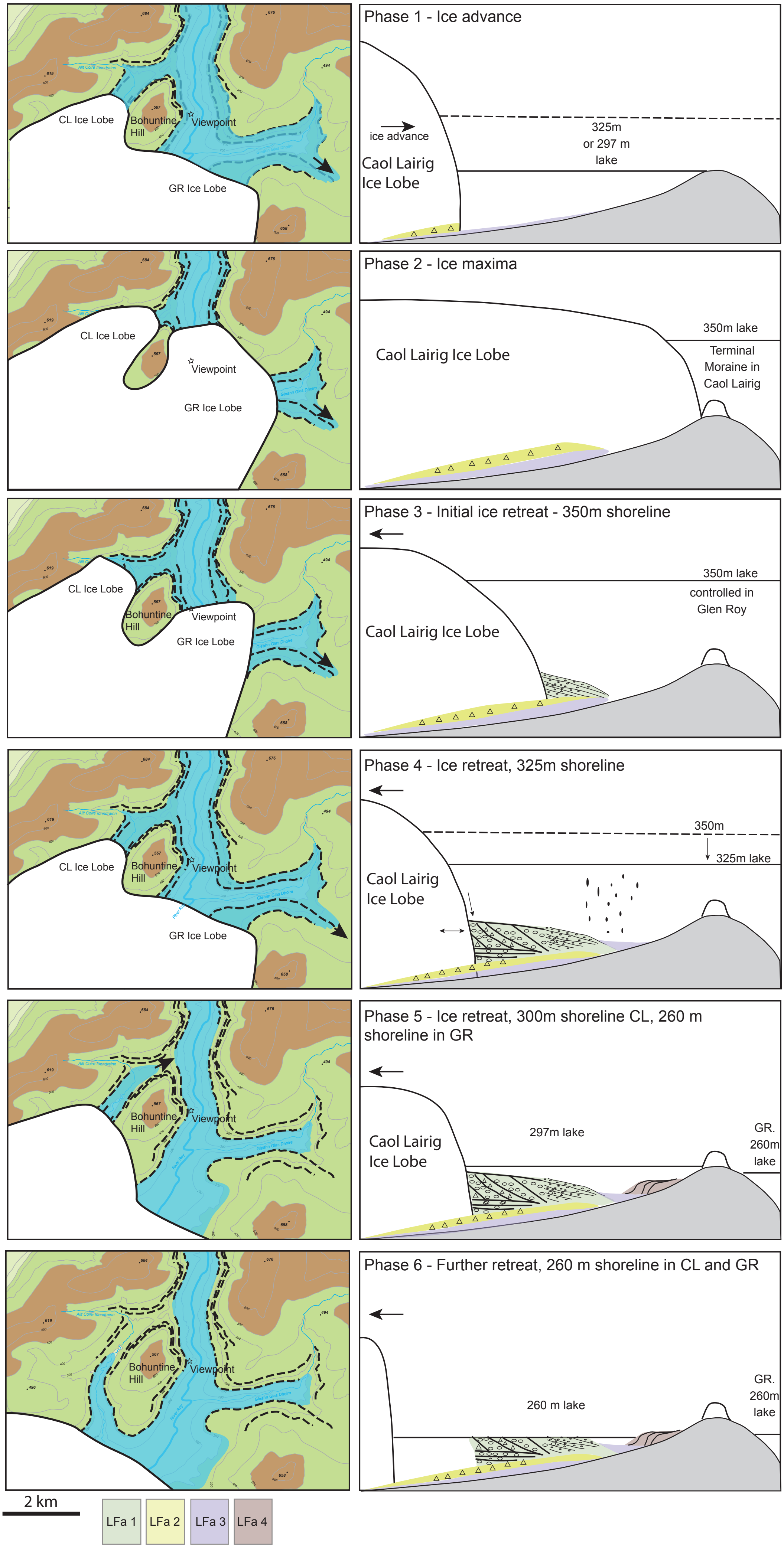

Key:
Lithofacies 1a Gravelly sand facies of proximal subaqueous fan deposits
Lithofacies 1b Gravel and diamicton facies of proximal subaqueous fan deposits
Lithofacies 2a Subglacial traction till
Lithofacies 2b Subaqueous debris flows
Lithofacies 3 Distal glaciolacustrine sediemntation
Lithofacies 4 Deltaic facies
Diamicton:
Dmm Diamicton, matrix supported, massive
---(s) sheared
Gravel:
Gms matrixsupported, massive
Gcm clast supported, massive
Gp planar cross bedded
Gfu upward fining
Gh horizontally bedded
Gsi matrix supported imbricated
Sands:
Sm massive
Sl horizontal and draped lamination
St medium to very coarse and trough cross bedded
Sfu fining upwards
Sd deformed bedding
Sr(s) ripple cross laminated (type S)
St(A) ripple cross laminated (type A)
St (B) ripple cross laminated (type B)
Silts and Clays:
Fl(v) Fine lamination with minor fine sand and very small ripples
Fl(v) - with rhythmites and varves
Fl(d) - with dropstones
Site CL1:
NN27861 85568
Site CL2a:
NN28012 85786
Site CL2b:
NN28102 85786
Site CL3a:
NN28344 85975
Site CL3b:
NN28296 86001
Site CL4:
NN 28457 86174
Site CL5:
NN 28356 86355
Site CL6:
NN28347 86232
Depth (m)
C Z S G C B
(246.63)
(243.13)
(259.67)
(254.07)
(273.29)
(263.69)
(274.89)
(271.49)
(278.08)
(265.83)
(274.96)
(271.56)
(279.78)
(275.78)
Gms
Dmm(s)
Sl
Gms
Gms/Sl/
Sm/Gfu/St(A)
Gms
Gh
Gms
Gms/Sh
Sl/Fl
Sm
Dmm
Gfu
Dmm
Gms
Gh
Gms
Sl/Fl
Gms
Gms/Sl
Fl(d)/Fp
Gms
Dmm
Sl/Fl/Fd
Fp(l)
Gms
Gh/Sh/Gsi
(Sfu/St(A)/Sr(s))
Gm
Sl/Fl(d)
Fl/Fl(d)
Sl/Fl
Gms
Gms
Dmm(S)
Fl
Sl/Fl/-
Fl(d)
St(A)
Gm
Dmm
Gms
CL1 Dmm LFA 2a
CL2b Gms LFA 1b
CL2a Dmm LFA 1b
CL3a Dmm LFA 2b
CL3a Gms LFA 1b
CL3a Gms LFA 1b
CL3b DmmLFA 2b
Palaeocurrents
- Ripples
Palaeocurrents
- Imbricated clasts
CL4 St LFA 4
CL4 Gms LFA 1b
CL4 Gm LFA 4
CL5 Gms LFA 2a
CL5 Dmm LFA 2a
CL6 Dmm LFA 2a

Figure 9

Figure 10



CL Ice Lobe
Bohuntine Hill
Viewpoint
GR Ice Lobe
Phase 1 - Ice advance
ice advance
Caol Lairig Ice Lobe
325m or 297 m lake
Phase 2 - Ice maxima
Caol Lairig Ice Lobe
350m lake
Terminal Moraine in Caol Lairig
Phase 3 - Initial ice retreat - 350m shoreline
Caol Lairig Ice Lobe
350m lake controlled in Glen Roy
Phase 4 - Ice retreat, 325m shoreline
350m
325m lake
Caol Lairig Ice Lobe
Phase 5 - Ice retreat, 300m shoreline CL, 260 m shoreline in GR
297m lake
Caol Lairig Ice Lobe
GR. 260m lake
Phase 6 - Further retreat, 260 m shoreline in CL and GR
260 m lake
GR. 260m lake
2 km
LFa 1
LFa 2
LFa 3
LFa 4